References

Browning, K. A. (1964) Airflow and precipitation trajectories within severe local storms that travel to the right of the mean wind. *J. Atmos. Sci.*, **21**, pp. 634–639

NSSFC (1980) Tornadoes: when, where, how often. *Weatherwise*, **32**, pp. 52–59

Rotunno, R. (1986) Tornadoes and tornadogenesis. In: Ray, P. S. (Ed.) *Mesoscale meteorology and forecasting*, American Meteorological Society, Boston, pp. 414–436

Back to basics:
Fog: Part 1 – Definitions and basic physics

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This is the first of four parts of a review of what is currently understood about the physics of the formation and dissipation of fog on land and at sea, and to what extent this understanding might inform and assist in forecasting fog.

Part 1 is devoted to introducing readers to some definitions and the basic physics required to discuss commonly occurring types of fog on land (Part 2) and at sea (Part 3), and relating this to the forecasting problem (Part 4). Sections with a more advanced scientific or technical content are in smaller print, and can be omitted by the reader as required.

Definitions
Visibility

Light from a distant object viewed by an observer is scattered (by dust particles and air molecules) out of the line of sight between the object and observer, while light from elsewhere gets scattered into the line of sight. This may reduce the brightness contrast between the object viewed and its background until the object can no longer be distinguished from its background.

The distance (between object and observer) at which this occurs is called the visibility, which (for a given atmosphere) depends to some extent on the level of background illumination and cannot be determined at night. A more accurate measure of visibility (used at some major airports) is the brightness of a steady source of light measured over a given distance. This measurement can be made night or day, and can be related roughly to an observer's (eye) estimate of visibility.

Fog

'Meteorological' fog occurs (by definition) whenever the horizontal visibility falls below 1 km; the terms 'mist' or 'haze' apply to visibilities in the range of one to a few kilometres. Cloud base descending to ground level is also experienced as fog. Fog consists of water droplets typically 10–20 micrometres (μm) in diameter. Fog droplets contain insoluble dust or smoke particles, or traces of dissolved salts (*e.g.* sodium chloride, ammonium sulphate) which formed the initial condensation nuclei on which the droplets condensed. Occasionally, fog is 'technically' reported (visibility < 1 km) in very polluted (smoky) atmospheres before significant droplet growth has commenced, and also in sandstorms or duststorms.

In water fogs, there is a rough relationship between visibility, the amount of fog water per unit volume, and mean drop size. A thick fog with a visibility of about 100 m will contain 0.1–0.2 $\mathrm{g\,m^{-3}}$ of fog water if the droplets are 10–20 μm in diameter.

Water vapour pressure

A gas exerts a force on any solid or liquid surface bounding the gas through (mainly elastic) collisions of the individual gas molecules with the surface. This force is conventionally expressed as pressure, which is force per unit area. The (highly variable) contribution of water vapour pressure to total atmospheric pressure in temperate latitudes is of the order of 1 per cent near the surface.

Dew point and saturation

The atmosphere at a given temperature and pressure can only hold a given (and sharply defined) amount of water vapour. If the actual amount of water vapour has reached this limit, the atmosphere is said to be saturated and the air temperature at its dew point. Saturated air overlying a liquid water surface (such as an ocean) will be in a state of equilibrium with such a surface – *i.e.* the rate of evaporation of water off the surface will be equal to the deposition rate of vapour back into the ocean.

Supersaturation

Further increase of vapour pressure and/or decrease of temperature in already saturated air normally leads to condensation of excess vapour into water on the surface as dew. Away from the surface, condensation cannot occur unless condensation nuclei are available for the excess water vapour to grow on as droplets. Without these nuclei the atmosphere would become supersaturated, which is an unstable state. In practice, condensation nuclei are always present but some degree of supersaturation is required to force droplet growth.

Saturation vapour pressure increases exponentially with temperature as shown in Fig. 1. If the water vapour pressure at a given temperature lies above the curve in Fig. 1 (*e.g.* at A), the atmosphere is said to be supersaturated. The degree of supersaturation is defined as the fractional excess of vapour pressure above its saturated value (*e.g.* the ratio AA′/A′B in Fig. 1) at the same temperature.

In fog, supersaturation is probably of the order of 0.01–0.03 per cent, but supersaturations of the order of 1 per cent may be attained in thunderstorm updraughts. Theory predicts that in a perfectly clean atmosphere a supersaturation of a few hundred per cent is required for water vapour molecules to form spontaneous clusters without immediate dispersion by molecular bombardment, and for these clusters to grow into a droplet. However, this condition cannot be reached even in the laboratory where large supersaturations can be reached by rapid decompression because (as was first found in the Wilson cloud chamber) ion tracks generated by natural radioactivity in the atmosphere act as condensation nuclei before water molecules can form their own spontaneous clusters.

Basic physics

The production of supersaturation in the atmosphere

Supersaturation in which fog or cloud droplets may form can be produced in the atmosphere in three ways:

(i) By ascent and resultant cooling of an air parcel (*e.g.* in a convective updraught associated with cumulus cloud formation).
(ii) By radiative heat loss.
(iii) By the mixing of two parcels of slightly unsaturated air initially having different temperatures.

The last (mixing) mechanism is important, and its operation can be seen by reference to Fig. 1. If two parcels of air with initial temperatures and vapour pressures defined by points C and D in Fig. 1 are mixed, the resultant mixture will lie at a point M somewhere along the line joining C and D, depending upon the relative masses of the initial parcels, and will be supersaturated provided CD intersects the saturated vapour pressure curve as shown. The resultant condensation of water vapour will release

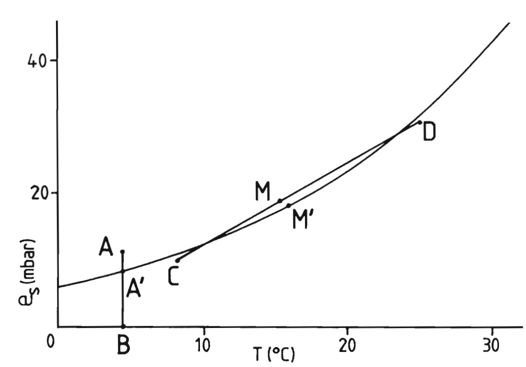

Fig. 1 Saturated water vapour pressure, e_s*, curve as a function of temperature,* T. *Other lines and symbols are explained in the text.*

latent heat, and the final mixture will end up near M′ on the curve. The maximum initial supersaturation that can be attained by mixing is roughly proportional to the square of the initial temperature difference between the air parcels which, in the atmosphere, has to be about 10 degC to generate a respectable fog.

The growth of fog droplets

The rate of growth of a fog droplet depends upon the temperature and the degree of supersaturation in the droplet environment, the nature of the nucleus on which the droplet is growing, and the droplet curvature. These factors control the flow of water vapour molecules and heat – latent heat is released by condensation – from and to the droplet surface. Droplet growth uses up water vapour and will therefore reduce local supersaturation unless this is being forced by cooling. Direct loss of infra-red radiation by a droplet can cool it and also have a significant (and sometimes dominant) effect on the rate of droplet growth.

In heavily polluted atmospheres, there is a large concentration of condensation nuclei. A result of this is that the fog water tends to be distributed between a large number of small droplets, rather than a small number of large droplets. Also, smaller droplets settle to the ground much more slowly, allowing a higher water content – a 'normal' fog initially about 30 m deep would deposit itself on the ground in about an hour if no other factors were present. It is thought that large numbers of small droplets combined with higher liquid water contents were characteristic of the old-fashioned 'pea soupers' (very rare nowadays) with visibilities of 10 m or less.

Heat transfer

The flow of heat across the Earth's surface (land or sea) and through the fog above plays an important rôle in fog evolution. The three main modes of heat transfer – radiation, convection and (molecular) conduction – all play their part.

The magnitude of heat flow (or flux) arriving at the Earth's surface does not change as it crosses the surface, although the mode of transfer does. This condition is known as the surface heat balance and is of fundamental importance in most meteorological studies. It states that heat flux conducted through the soil to the surface is equal to the heat removed from the surface into the atmosphere by radiation and turbulent diffusion, *i.e.* stirring/mixing by wind. This also applies for heat flowing from the atmosphere to the soil, as happens in daytime.

The surface heat balance is usually written in the form:

$$G = R + C + E.$$

The sign convention is that heat fluxes directed downwards are positive. G is the soil heat flux at the surface, R is the net radiative flux at the surface, while C and E are the fluxes of sensible (C) and latent (E) heat transferred by turbulent diffusion. In other words, C represents warm air convected away from the surface, and E represents the latent heat required to evaporate water from the surface so it can be taken into the atmosphere by turbulence.

Figure 2(a) shows a schematic representation of the surface heat balance on a clear night with light winds, with the near-surface heat flux shown in Fig. 2(b). The loss of radiation from the surface is partly offset by weak turbulent diffusion of sensible and latent heat to the ground. In order to satisfy the heat balance, the sum of the R, C and E arrows equals the soil heat flux arrow G. According to the sign convention, R and G are negative quantities, while C and E are positive. On a sunny day, all fluxes are much larger and in the opposite direction – Fig. 2(c).

The radiation term at the surface consists of four parts: downward fluxes due to (i) incoming solar radiation and (ii) infra-red (heat) radiation emitted from cloud, water vapour and other gases, and upward fluxes due to (iii) reflection of solar radiation at the surface and (iv) infra-red radiation emitted from the surface.

Nearly all the energy of solar radiation lies between wavelengths of 0.4 and 4 μm, whereas nearly all the infra-red radiation energy lies between about 4 and 50 μm. The atmosphere lets this radiation through in some wavelengths but not others. The main transparent regions (or 'windows') of the atmosphere are in visible wavelengths (0.4–0.7 μm) and in the middle of the infra-red range (8–13 μm). At other wavelengths, radiation is absorbed by minor gaseous constituents of the atmosphere – mainly water vapour and carbon dioxide – but not by the

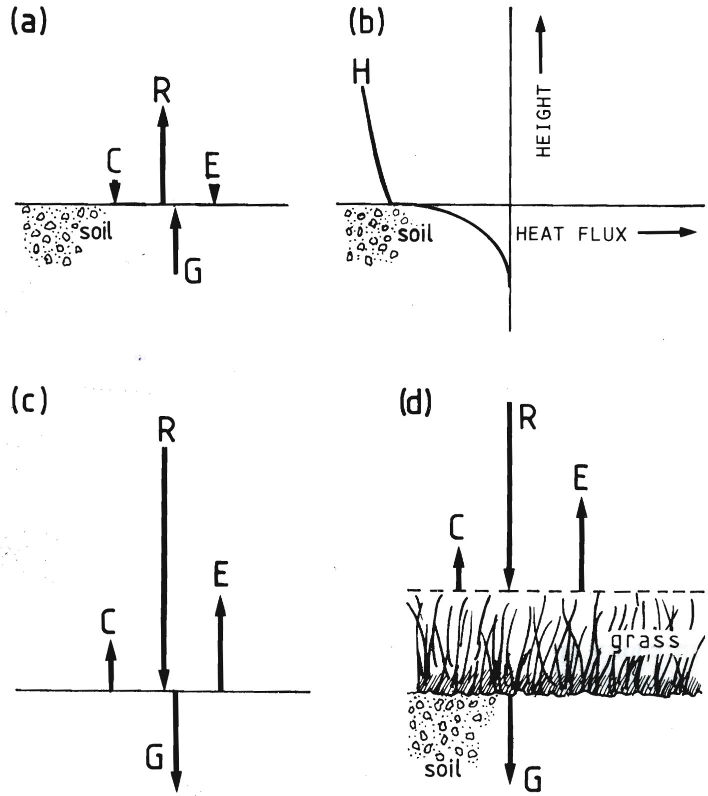

Fig. 2 Schematic representation of (a) surface heat balance on a clear night and (b) total heat flux (H) near the surface, also on a clear night. (c) represents the surface heat flux balance on a sunny day (R is now the incoming solar radiation minus a smaller (upward) infra-red component), while (d) is the surface heat balance when the soil is covered by a layer of grass. (NB Total heat flux (H) is the sum of R + C + E above the surface, and G below the surface.)

main gases, nitrogen and oxygen.

All solid bodies emit and absorb infra-red radiation over a wide wavelength range, but water vapour and carbon dioxide only do this in limited wavelength bands. The amount of energy at each wavelength depends on temperature, and can be calculated using the quantum theory of radiation. The result of this is that the ground radiates energy to the atmosphere over the wavelength band 4–50 μm. Some of this radiation escapes to space through the 8–13 μm window, but some is absorbed by water vapour and carbon dioxide, which emits radiation back to Earth. Under a cloudless atmosphere, the surface will receive back from the atmosphere typically 70–80 per cent of the radiation it emits; under a layer of cloud or fog, this fraction rises to 90–100 per cent because the water in the cloud absorbs infra-red radiation. This 'blanket' effect of the atmosphere and clouds is often referred to as the 'greenhouse' effect.

There are two important complications in the handover of heat from the surface to the atmosphere (and vice versa).

(i) While radiation takes place directly from the surface, the heat removed by turbulence has first to be conducted through a thin skin of air in contact with the surface before turbulent diffusion can 'get hold' of it. Since air is a very poor conductor of heat, temperature differences of several degrees between the surface and the air a few millimetres above may occur in order to satisfy the surface heat balance. This is why frost sometimes forms on the ground under a clear night sky, although the minimum air temperature at screen level (1.2 m above the ground) may not fall below 3–4 °C. It is also why, on a hot summer's day, dry sand may be too hot to stand on with bare feet as it can be 20–30 degC above air temperature.

(ii) The surface heat balance over land is complicated by the presence of vegetation. In this case, the handover from pure molecular conduction to radiation, etc. is smeared over the depth of the vegetative canopy. This also tends to reduce the large temperature differences between surface and atmosphere occurring over bare ground, although these can still be quite large over short turf. The surface heat balance when grass cover exists is shown in Fig. 2(d).

The presence of vegetation raises the problem of where the level of the surface is defined. However, heat storage in the canopy is small compared with the uncertainties in estimating the main terms, so that the soil heat flux at the base of the canopy will differ little from the sum of radiation, convection and evaporation at the top of the canopy.

The fundamental difference between land and sea surfaces as regards heat exchange is that land surface temperature is *controlled by* the surface heat balance, whereas sea surface temperature (SST) *controls* the surface heat balance. Whereas the diurnal range of land surface temperatures may exceed 30 degC, the diurnal range of SST is very small, so that SST can be considered constant (with time but not distance) over a day or so. It is not necessary for short-term meteorological purposes (but very important in climate studies) to consider the flux of heat below the sea surface. Turbulent sensible/latent heat fluxes across the sea surface are expressed in terms of the wind speed and temperature/vapour pressure differences between sea surface and atmosphere near the surface – the so-called 'bulk aerodynamic' relations.

Conclusion

The basic physics outlined above is well established. The mechanisms which generate supersaturation, and the way fog droplets grow in a supersaturated atmosphere are also well understood.

What has *not* been understood is how the processes generating supersaturation interact with other physical and dynamical processes (*e.g.* radiative transfer, turbulence, droplet settling, wind field, fog-top entrainment, etc.) in the generation and dissipation of land fog and sea fog in different meteorological situations. Recent studies of this problem are discussed in Parts 2 and 3.

Two memorable winters in Scotland – 1793/94 and 1794/95:
'The Gonial Blast' and 'the most severe in the memory of man'

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Winters in temperate latitudes may become notorious either because of a single devastating event or on account of a prolonged battering by the elements. Two such winters occurred in succeeding years in Scotland: during 1793/94 in the south-west and, most severely in the east, during 1794/95.

'The Gonial Blast'

As a curtain-raiser to the prolonged winter of 1794/95 a remarkable snowstorm swept the south-west of Scotland in January 1794. It came to be known locally as 'The Gonial Blast' because of the extraordinary number of sheep